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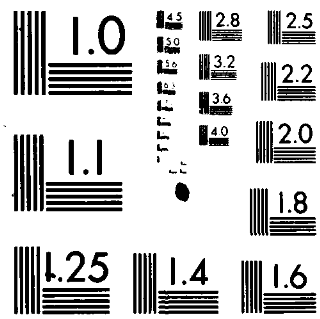

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OCEAN/TROPOSPHERE/STRATOSPHERE FORECAST SYSTEMS: A STATE-OF-THE-ART REVIEW

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The question of the feasibility of developing a coupled Ocean-Troposphere-Stratosphere Forecast System is discussed. The system would be run operationally by Fleet Numerical Oceanography Center (FNOC). Specifically, the state-of-the-art of ocean-atmosphere modeling is reviewed to decide if such a forecast system is feasible during the next five years (1982-87).

The conclusion of the review is that such a system is not operationally feasible in the next five years. (Continued on reverse)

Abstract, Block 20, continued

Modeling research efforts at various institutions and laboratories have been unable to reproduce many of the observed interactions between the ocean-troposphere and troposphere-stratosphere. In some instances, models have produced erroneous interactions, actually degrading atmospheric forecast skill when compared to models without these interactions. Until such time as the state-of-the-art of coupled ocean and atmosphere models can guarantee benefits from simulating these interactions, operational use of them is not advised.

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SECTION 1 - SUMMARY

The state-of-the-art of global numerical forecast models of the ocean and atmosphere has been reviewed and is presented in Section 2 of this report. The primary aim of this review was to determine if the accuracy of a medium-range operational atmospheric forecast model could be improved during the next five years through the coupling with a global ocean model and/or the inclusion of stratospheric modeling within the atmospheric prediction model. The primary conclusion of this review is that accuracy could not be improved through those means. There is evidence to show that inclusion of an interactive ocean model, with the present accuracy of such prediction models, may deteriorate rather than improve the atmospheric prediction accuracy. The evidence for stratospheric interaction indicates little, if any, effect on the troposphere. Another important conclusion is that some of the most relevant atmospheric/ocean interaction on a time scale of up to 10 days occurs on the regional scale. Large sea surface temperature gradients along ocean fronts and/or eddies and water/ice boundaries in polar regions can be responsible for rapid extratropical cyclogenesis.

This review does not pertain to ocean models which are to support antisubmarine warfare operations or stratospheric modeling to improve stratospheric forecasts without a subsequent improvement within the troposphere.

The recommendations of this extensive study are presented in Section 3 of this report. The main point of these recommendations are the following:

(1) Do not add detailed stratospheric modeling to the global atmospheric forecast model,

(2) Do not couple the global atmospheric forecast model with a global ocean forecast model,

(3) Increase the vertical resolution of the global

atmospheric forecast model to about 12 layers with the upper forecast level at 50 mb and with an increase in the lower tropospheric resolution above the boundary layer to predict properly the response to sea surface temperature gradients in regions of strong air-sea interaction,

(4) Increase the horizontal resolution of the global atmospheric forecast model to about 2°-by-2° in latitude and longitude,

(5) Improve the modeling of diabatic processes in the troposphere, and especially the atmospheric response to ocean interaction in regions of large fluxes,

(6) Develop a global ocean analysis-prediction system to describe accurately the initial conditions and produce reasonable forecasts of surface parameters to the time of the next oceanic operational analysis cycle (12 or 24 hours),

(7) Upgrade the Cyber 200 model 203 computer to a model 205 with at least 2 million words of central memory, and

(8) Acquire CDC 32-bit Fortran compiler.

Associated recommendations for the most cost effective expenditure of funds to improve the medium-range forecast capabilities of the operational atmospheric forecast model executing at the Fleet Numerical Oceanography Center (FNOC) during the next five years are not summarized. These recommendations should be read in context with the review material presented in Section 2 of this report.

SECTION 2 - STATE-OF-THE-ART OF GLOBAL ENVIRONMENTAL NUMERICAL FORECAST MODELS

2.1 Introduction. This section is presented in three parts. The first part reviews the state-of-the-art of global ocean prediction modeling relative to the requirements of medium range atmospheric forecast models. The second part presents the state-of-the-art of coupling global ocean prediction models with global atmospheric prediction models. The third and last part of this section reviews the state-of-the-art of including stratospheric modeling within global atmospheric prediction models. The depth and thrust of all these reviews is relative to improving medium range forecasts of the atmosphere within the troposphere. Thus, the objectives of the ocean and stratospheric modeling reviews are not to demonstrate how forecasts within these regions may be improved without an associated improvement within the troposphere. This review is restricted to the question of ocean-troposphere-stratosphere interactive global modeling. The important questions of atmospheric analysis and data assimilation and global model formulation are not addressed.

2.2 Global Ocean Forecast Models.

2.2.1 Background. As described at a recent workshop on ocean prediction (Mooers, *et al.*, 1981), a complete ocean description-prediction system consists of three main components: (1) a real-time data collection system; (2) analysis schemes to characterize the present state of the ocean; and (3) various models to forecast the future state of the ocean. The Navy has needs for ocean prediction on a variety of scales and in a variety of geographical regions. Each of these regions may present a unique oceanographical prediction problem. For example, the tropical and equatorial regions, the dynamically active western boundary regions, the marginal ice zones, the midlatitude open oceans and the eastern boundary upwelling regions all require somewhat special treatments because of unique features or processes that dominate the physical oceanography of the particular region. Thus, the Navy's needs for ocean predictions may best be served by having a unique prediction strategy (observations, analysis and models) for each specific prediction need or each particular geographical region. Because this review addresses the specific question of ocean-atmosphere model coupling, it will only be concerned with ocean prediction systems that are relevant to sea surface temperature (SST) predictions on a scale which affects the atmospheric synoptic scale during a 10-day forecast period. Thus, this review of the state-of-the-art of ocean prediction is limited to the larger synoptic scales of ocean variability, covering the global oceans, that are addressed by prediction systems such as the Navy Operational Global Atmospheric Prediction System (NOGAPS) and Thermal Ocean Prediction System (TOPS). The review will cover first the status of operational global analysis (description) and then the status of synoptic ocean modeling (prediction). In the analysis category, there are essentially two operational systems - that of the FNOC and that of the National Oceanic and Atmospheric Administration's (NOAA) National Weather Service.

2.2.2 Ocean Analysis. At present, FNOC's daily ocean thermal structure analysis is based on approximately 200 XBT's and 3000 surface observations (ships and buoys) over the whole globe (Petit, 1981). These limited number of conventional observations are furthermore very unevenly distributed. Satellite data are not used. The analysis scheme is a "successive corrections" type of objective analysis which contains no explicitly modeled physics and is based entirely on standard information blending techniques (Mendenhall *et al*., 1978; Holl *et al*., 1979). Information is blended vertically as well as horizontally, so the SST observations contribute to the subsurface thermal analysis. Since the first guess field is a "forecast" of adjustment toward climatology, the analyzed fields in data-sparse regions remain near the monthly climatology. This heavy reliance on climatological data is expected to be discontinued when the TOPS model is introduced operationally at FNOC. At that time, the first guess field will probably be provided by the model prediction in much the same way as is done in operational analysis/forecast systems in meteorology (McPherson et al., 1979; Elsberry and Garwood, 1980; Clancy and Martin, 1981).

The operational analyses prepared by NOAA consist entirely of ocean surface properties and they are primarily for regional, not global, areas (U.S. Department of Commerce, 1980). Two manually prepared "regional oceanographic analyses" and two objectively analyzed SST charts are transmitted regularly on the NAFAX and DIFAX facsimile circuits. The regional oceanographic analyses are prepared manually by a joint team from the National Weather Service (NWS) and the National Environmental Satellite Service (NESS). Satellite imagery with a resolution of 1 km is used to locate ocean features such as the Gulf Stream, slope fronts, loop current and eddies, while conventional data (ships, buoys, XBT's) together with 50 km satellite retrievals are used to determine temperatures associated with various features and water masses. The objectively analyzed SST charts are prepared using conventional data composited over a five-day period. A first guess of the SST field (taken from the previous five-day analysis or from climatology) is adjusted with these observations using a form of the Cressman objective analysis technique in which acceptable observations within a prescribed radius of influence (weighted by distance, not time) around each (25 km) grid point are used to modify the first guess field. No ocean physics is explicitly modeled. The final field is smoothed and contoured from a 50 km grid. These SST analyses are prepared each week for the Northwest Atlantic/Gulf of Mexico region and for the Gulf of Alaska/Eastern Pacific region.

NOAA also prepares monthly mean SST analyses on hemispheric and regional grids based on conventional and satellite (GOSSTCOMP) data (Brower et al., 1976; U.S. Department of Commerce, 1981). Although individual satellite temperature retrievals may not presently have the accuracy required by most Navy needs (Barnett et al., 1979), where one is able to get a sufficient density of observations over an extended period per unit area, a more credible temperature estimate can be achieved (Strong and Pritchard, 1980). The conventional and satellite data are used to prepare two-day composite hemispheric SST analyses on alternate days for the Northern and Southern Hemispheres. The data are analyzed on polar stereographic grids having a mesh length of 190.5 km at 60° latitude using an objective analysis technique known as "conditional relaxation". This technique corrects the previous analysis with those new observations which have passed a variety of validity checks. Areas without observations are modified by the "relaxation" procedure which maintains the divergence of the SST gradient in these areas. Each two-day hemispheric analysis is interpolated to a 2° latitude/longitude grid and monthly means and monthly anomalies [relative to the Robinson-Bauer climatology (Robinson, 1976; Robinson et al., 1979)] are computed. A similar procedure is followed in preparing the monthly mean SST analyses on regional grids except that the data are composited over five days, the polar stereographic grid mesh is 23.8 km at 60°N, and the Cressman objective analysis technique is used. The five-day composite analyses are prepared twice a week and then interpolated to a 1° latitude/longitude grid to compute the monthly means.

The above state-of-the-art of operational ocean analysis can be summarized as follows. The Navy's system consists entirely of blending a generally inadequate number of in situ observations with a climatological data base, while NOAA's approach makes additional use of satellite retrievals. These are used to portray the details of ocean thermal features such as fronts, eddies, upwellings, etc., where absolute temperature values are not critical (Richardson, et al., 1973; Stumpf and Legeckis, 1978; U.S. Department of Commerce, 1980), and to prepare objective analyses of monthly mean SST where time-compositing effectively removes many of the day-to-day (random) errors caused by clouds and water vapor (Strong and Pritchard, 1980; U.S. Department of Commerce, 1981). Unfortunately, because of the current paucity of conventional observations and the relatively large errors associated with individual satellite retrievals of SST, the uncertainties in describing the ocean initial state on a daily basis are as large as (if not larger than) the errors expected in numerical model predictions due to an imperfect representation of model physics (Clancy and Martin, 1981; also see below).

In the near future, three major areas of basic and applied research are likely to result in substantial improvements in daily ocean analyses. The first is the implementation and utilization of a number of important satellite sensor improvements and analysis methods for retrieving the SST over the global oceans (McClain, 1980; Deepak, 1980). The improved sensors include the Advanced Very High Resolution Radiometer (AVHRR) and the multi-channel High resolution Infrared Radiation Sounder (HIRS) which are now available on NOAA satellites (Schwalb, 1978; Stewart, 1981). These sensor improvements were incorporated into NOAA SST analyses on November 17, 1981 (Pichel, 1981). New advanced methods of analysis such as the split window technique in the infrared to remove water vapor effects by Chahine, McMillan, Crosby and DePriest, and the use of truncated normal radiance distributions to remove cloud effects by Fleming (Deepak, 1980) should also become operationally feasible in the near future. As a result, it now appears that the goal of achieving global SST distributions with accuracies of 1°C at resolutions of a few tens of kilometers from satellite radiances alone is within reach (Leovy, 1981). In addition to SST, sufficiently reliable estimates of the global surface wind field may eventually be operationally feasible using data from the Scanning Multichannel Microwave Radiometer (SMMR) and the radar Scatterometer (SCAT) (O'Brien, 1981; Stewart, 1981). As noted below, such global surface wind and SST analyses will greatly increase the utility of upper ocean models in describing the oceanic initial state. Ultimately, in the more distant future, global measurements of the surface topography of the ocean may be available on an operational basis using radar altimeter measurements from satellites (Wunch, 1981). Such observations of the sea surface elevation would make it possible to initialize a low vertical resolution dynamical model having fine (∿50 km) horizontal resolution. This model would be capable of providing a reasonable description of oceanic synoptic (eddy) scale features. Even with such fine horizontal resolution, useful model prediction out to 10 days could be made once a week with present computers. While the potential use of satellite data in ocean analysis is obviously very great, it is clear that progress and full operational realization will require a patient and dedicated collaboration among space engineers, data processing specialists and oceanographic scientists (Bretherton, 1981; Goody, 1981).

The second major development which is expected to improve the initial thermal structure analysis is the use of a modern objective analysis scheme to make optimum use of the available observations (White and Bernstein, 1979; Williams et al., 1981). Objective analysis schemes make use of important observed statistical quantities, such as the local space and time scales of variability, to help map the fields in areas of sparse data. These observational statistics are sufficiently well known for this purpose in many geographical areas (Bernstein and White, 1974; Dantzler, 1976; White and Bernstein, 1979; Barnett and Patzert, 1980; White et al., 1981). The use of a modern objective analysis scheme in the ocean is expected to make a significant improvement in the ocean thermal structure analysis below the mixed layer, where the observations are extremely sparse and the statistical properties quite different from those of the atmosphere.

The third major research development which will greatly improve the initial analysis (description) of the ocean thermal structure is the use of a physically consistent ocean prediction model in the daily analysis-prediction-analysis cycle (Elsberry and Garwood, 1980). Such data assimilation methods are widely used in operational Numerical Weather Prediction (Haltiner and Williams, 1980). Quite recently, the use of models to generate consistent and improved initial conditions for ocean predictions have been tested (Robinson and Haidvogel, 1980; Warrenfeltz, 1980). While the cases considered were very idealized, the results successfully demonstrate the use of models in oceanic data assimilation. In the technique used by Warrenfeltz, the mixed-layer model is used to advance (i.e. hindcast) the observations forward in time to the initial time of a forecast. If a sufficiently large number of observations are available during the 30-day period prior to initial time, a procedure for removing random observational errors was demonstrated. As noted above, however, it is the quantity rather than the quality of observations which is so low in the ocean. This topic is further examined in the context of the NOGAPS-TOPS system in subsection 2.3.

2.2.3 Ocean Prediction Models. In the ocean description-prediction system of the future, the ocean model will clearly play a central role. As noted above, the model's first and perhaps most important role will be to assimilate (i.e., to spread, integrate and filter) the available observations into a dynamically consistent initial description of the ocean. No model can perform this function adequately without reliable estimates of the atmospheric forcing. It has been demonstrated that short term variability and peaks in the atmospheric forcing are very significant to ocean prediction (Camp and Elsberry, 1978; Elsberry and Garwood, 1978; Price *et al.*, 1978; Klein, 1980; Clancy and Martin, 1981; Krauss, 1981; Klein and Coantic, 1981; and many others). Due to insufficient meteorological data over the ocean, coarse horizontal resolution, and the use of smoothing operators in the objective analysis (blending) schemes, the present analyses of the atmospheric forcing fields over the ocean are generally not adequate for ocean prediction models (Elsberry and Garwood, 1980). The data acquisition and analysis schemes used in the future must be capable of extracting these important (sub)synoptic scale events from the available observations. The relatively high resolution global SST and surface wind data that can be obtained from satellites therefore appear to be essential for the future of ocean description and prediction. However, it should be noted that since the SST is not the upper boundary condition for ocean models, and since the fluxes of heat and moisture across the sea surface can not presently be measured from satellites, major problems still remain in using models and remotely sensed data for even the comparatively simple role of ocean description.

At the present time, the use of a dynamical forecast model for the global ocean is not feasible. There are two reasons for this. First, the oceanic synoptic scale variability, i.e. that part of the variability which is potentially predictable with such a dynamical model, occurs on a much smaller space scale ($\sim$100 km) than is resolved by the present global grids. Thus, even if the oceanic subsurface synoptic (eddy) fields were adequately observed (say on a 50 km scale by satellites) the large-amplitude, small-scale features would be seriously aliased into the larger scales that are resolved by the grids. The second reason why a dynamical model is not appropriate for ocean prediction at the present time is because the present subsurface observational network is not adequate. As noted above, relatively high resolution altimetry data, when available, would provide a means of initializing the lowest vertical modes in such a dynamical model if the horizontal resolution were adequate. At the present time however, the introduction of extremely sparse subsurface data into a (relatively) coarse grid dynamical model would simply result in spurious eddies having scales determined by the influence function of the analysis scheme.

Ocean prediction models appropriate for the space and time scales addressed by the NOGAPS/TOPS system are the class of models commonly referred to as "mixed-layer" models (see Garwood (1979) for an excellent review). The two basic assumptions in the mixed-layer models are (1) that vertical mixing in the upper ocean occurs as a result of local atmospheric forcing and (2) that the key to obtaining closure is in the mechanical energy equation. As a result, mixed-layer models are usually classified according to the assumed form of the model or the method of attaining closure.

One large class of mixed-layer models is that known as "slab (integral) models" (Kraus and Turner, 1967; Denman and Miyake, 1973; Pollard et al., 1973; Niiler, 1975; DeSzoeke and Rhines, 1976; Elsberry et al., 1976; Gill and Turner, 1976; Kim, 1976; Thompson, 1976; Garwood, 1977; Pollard, 1977; Price et al., 1978; DeSzoeke, 1980). In these models, the temperature, salinity and horizontal velocity (if included) are assumed to be quasi-uniform within a "well-mixed layer" (see Niiler and Kraus, 1977; and Zilitinkevitch et al., 1979 for reviews). The slab models are based on assumptions about the integral effects of several turbulence mechanisms (i.e. production, dissipation and transport), of which there is still considerable uncertainty. The major source of contention among the slab models is whether entrainment mixing at the base of the well mixed layer is accomplished primarily by a mean (shear) flow instability (Pollard et al., 1973; Thompson, 1976; Pollard, 1977; Price et al., 1978) or by a convergence of flux of turbulent kinetic energy generated near the surface (Kraus and Turner, 1967; Denman and Miyake, 1973; Elsberry et al., 1976; Kim, 1976; Garwood, 1977). Many recent mixed-layer models contain both mechanisms (Niiler, 1975; DeSzoeke and Rhines, 1976; Price et al., 1978; DeSzoeke, 1980; Davis et al., 1981; Adamec et al., 1981). The question of the relative importance of the two different entrainment mechanisms is not simply an academic question. Those models in which the entrainment is parameterized primarily in terms of mean shear are extremely sensitive to the surface stress forcing because the mean shear is largely due to vigorous inertial motions induced by the working of the wind. Such models are therefore sensitive to the duration of the forcing relative to the local inertial period and to the direction of the forcing relative to the existing inertial motions (Krauss, 1981; Price, 1981). Models in which the entrainment is due to turbulence produced by surface shear are not sensitive to the surface wind direction (the wind forcing enters only as u_*^3, where u_* is the friction velocity) nor to any residual inertial motion in the mixed layer (Elsberry and Garwood, 1980).

The second major class of ocean mixed-layer models is the "second order closure" models (Mellor and Yamada, 1974; Wyngaard, 1975; Wyngaard and Cote, 1975). The TOPS model is of this type (Clancy and Martin, 1981). In most oceanic applications of these models, the vertical turbulent fluxes ($\overline{w'u'}$, $\overline{w'\theta'}$ etc.) are assumed to be related to the mean gradients ($\partial\overline{u}/\partial z$, $\partial\overline{\theta}/\partial z$ etc.) in the traditional K-theory format (Mellor and Durbin, 1975; Klein, 1980; Klein and Coantic, 1981; Clancy and Martin, 1981). The turbulent vertical eddy diffusivities are obtained from steady state forms of the second order turbulence equations and thereby depend directly on the local current (vertical) shear and dynamic stability. The resulting vertical eddy diffusion coefficients are continuous functions of the dynamic stability and the current shear. They become zero, implying a cutoff of turbulent mixing, for values of the gradient Richardson number greater than 0.23 (a number determined from laboratory experiments). As pointed out by Garwood (1979), the vertical flux of turbulent kinetic energy is neglected in the derivation of these closure schemes. Thus, as shown by Martin (1976), these closure models behave more like the slab models that are based on mean shear (Pollard et al., 1973) than those that are based on surface stirring (Kraus and Turner, 1967). Consequently, these models are also very sensitive to the characteristics of the surface wind forcing (Mellor and Durbin, 1975; Klein, 1980; Klein and Coantic, 1981; Clancy et al., 1981).

At the present time, most mixed-layer models have only been tested against data collected at the Ocean Weather Ships (Thompson, 1976; Camp and Elsberry, 1978; Elsberry and Raney, 1978; Elsberry and Garwood, 1980) or data collected as part of special process-oriented field programs (Price et al., 1978; Davis et al., 1981; Krauss, 1981; Klein and Coantic, 1981; Warn-Varnas et al., 1981). The TOPS model has been more widely tested using the FNOC operational data base (Clancy and Martin, 1981; Clancy et al., 1981). From the above studies, a general awareness of the level of forecast skill attainable with such models is slowly emerging.

For example, during the preliminary FNOC test and evaluation of TOPS, a sequence of 37 daily three-day forecasts was performed with the non-advective version of the model using operational fields during Fall 1980 (Clancy et al., 1981). The thermal field was initialized from the daily Extended Ocean Thermal Structure (EOTS) analysis and verified against the EOTS analysis valid at the end of each forecast period. The resulting pattern correlation between the forecast and analyzed SST changes for each three day period fluctuated between 0. and 0.5 with an average of 0.25. This demonstrates a potentially useful level of forecast skill in a realistic operational setting.

It is not clear whether any of the other models reviewed above would have performed better in a similar test. The Mellor and Durbin model (the ancestor of the TOPS model) was also used by Klein (1980) in an attempt to simulate the marine upper layers in the Gulf of Lion during a special COFRASOV II expedition in July 1976. The model was relatively successful in simulating the SST and mixed layer depth changes but only if the atmospheric drag coefficient was given an unrealistically high value. Klein (1980) noted the extreme sensitivity of the model to the wind forcing (see also above comments), and the need to use a large coefficient. Perhaps the most successful model simulation, having the advantage of special observations both for forcing and verification, is the simulation of the Mixed-Layer Experiment (MILE) data by Davis et.al. (1981). They used the Niiler (1975) slab model which contains a parameterization of entrainment due to both surface stirring and mean shear instability. They found that the potential energy changes in the water column were highly correlated with u_*^3 and that mean shear instability was only important at the onset of wind events. This result is consistent with the analysis of DeSzoeke and Rhines (1976) who also showed that there is a severely restricted depth below which the mean shear mechanism cannot mix (unless, apparently, C_D is made artificially large). With a judicious choice of two parameters ("tuning" the two mixing mechanisms), and by taking vertical advection into account below the mixed layer, Davis et al. (1981) have been able to replicate very closely both the SST and the mixed layer depth evolutions during MILE. The surprising result to them, however, was that the observed vertical and temporal structure of the horizontal currents was reproduced as well. The authors also considered it very significant that the fundamental "slab" assumption was apparently verified. In summary, it appears that a state-of-the-art mixed layer model which contains the proper balance of the dominant mixing mechanisms in the upper ocean can provide very useful simulations of SST and mixed layer depth, given the proper forcing. These issues are discussed further in the next subsection on coupled atmosphere-ocean prediction models.

2.3 Coupled Atmosphere-Ocean Forecast Models.

2.3.1 Atmospheric Response to Ocean Heat Flux. A schematic illustration of the effect that the ocean has on atmospheric flows is given in Exhibit 1. The abscissa represents a normalized measure of the influence of the ocean and its time changes on the local state of the atmosphere. This influence varies in different ocean regions and in different seasons - due to changes in the internal structure (stability) of the atmosphere and the ocean. This exhibit illustrates that the initial distribution of atmospheric properties determines the future evolution of the atmosphere during the first 2-3 days of a forecast, with the ocean's effect becoming more important at longer time scales. One illustration of these trends was provided by Kraus and Morrison (1966). They showed that the air temperature was the primary contributor to the air-sea temperature difference on time scales of a few days. However, the ocean temperature variations contribute more to the variance in the air-sea temperature difference on monthly and longer time scales. Between the two extremes, the effect of the ocean on the atmosphere must become progressively larger with time. It is the extension of the atmospheric forecasts into the medium ranges (5-15 days) that makes the question of the relative importance of ocean effects an important one.

Suppose that one has a perfect atmosphere model and a perfect ocean model. As shown in Exhibit 2, the coupled version of these perfect models would have a skill of 1.0 with increasing time. If the trend shown in Exhibit 1 is correct, integrating the perfect atmospheric model uncoupled from the ocean model would result in an increasingly large reduction in skill with time. The atmospheric model skill would be reduced to near zero on the seasonal time scale. Of course, there are no perfect models and the data are not adequate to specify perfectly the initial conditions in the atmosphere or in the ocean. For the case without an ocean, the decrease in skill (arrow "b") with an imperfect atmospheric model is thus larger than shown for the perfect model in Exhibit 2. Including a perfect specification of the ocean conditions would recover some of this decrease in skill. However, coupling the atmospheric

EXHIBIT 1. Schematic illustration of how the oceans effect on atmosphere (on "synoptic" scales) increases with time of forecast.

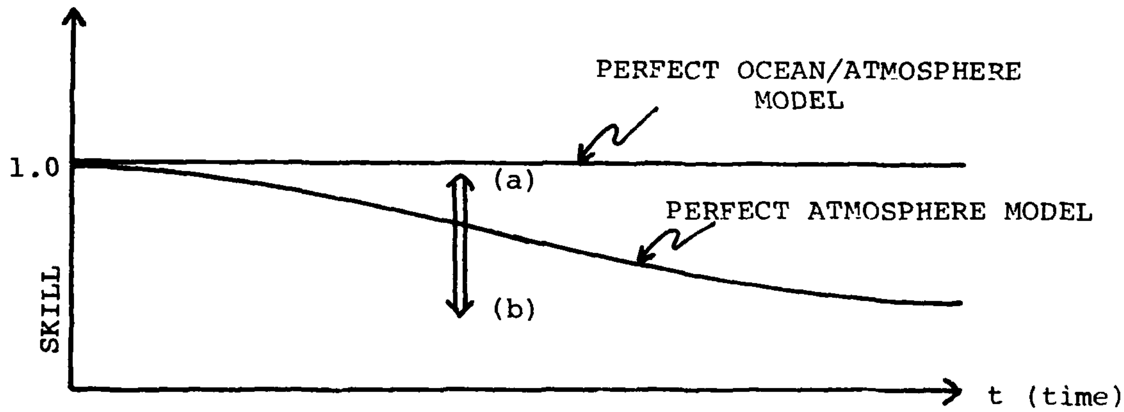


EXHIBIT 2. Schematic illustration of how the skill of a perfect atmosphere model will decrease with time due to the absence of ocean effects. Since the effect of the ocean on atmosphere is small at times less than 10 days, introducing a large uncertainty (degrees of freedom) in troposphere due to uncertainties in the ocean model and in coupling may cause the skill to decrease.

model with an imperfect ocean model also introduces additional degrees of freedom into the atmosphere. One must at least consider the possibility of enhanced growth of errors in the atmospheric model. Furthermore, it is possible that any increase in skill due to the inclusion of the effects of the ocean (arrow "a") may be small compared to decrease in skill due to error growth in the atmospheric model.

The regions and seasons during which the ocean changes are likely to affect medium-range atmospheric predictions are relatively well known. The most sensitive region is clearly in the equatorial areas. Changes in sea surface temperature of a few degrees in the equatorial regions contribute to deep convection in the atmosphere, because of the conditionally stable thermal structure in the lower troposphere. Whether these changes are correct or incorrect, the effect is felt in a deep tropospheric layer on time scales of less than a day. The atmospheric response is not just felt locally. An extensive literature (see recent papers by Hoskins and Karoly, 1981; Webster, 1981) has developed regarding the effect of warm equatorial anomalies. There appears to be good evidence that the effect of these anomalies can be felt in the long waves of the extratropics within seven days after the imposition of the anomaly. The conditions and mechanisms by which the effects of equatorial influences are felt remotely are being studied by many groups.

The western boundary regions of the mid-latitude oceans are prime areas for cyclogenesis. Warm ocean currents, such as the Kuroshio and the Gulf Stream, extend poleward along the east coasts of the continents. As these warm currents interact with cold, equatorward currents, large horizontal gradients of sea surface temperature are formed. It is clear that these coastal and open-ocean regions with large horizontal gradients of surface air temperature are important in cyclogenesis. The relative contributions of the magnitude versus the precise location of the sea surface temperature gradients is less clear.

It may be considerably more difficult to predict accurately the locations of mesoscale ocean eddies than it will be to predict the modification of the sea surface temperature gradients across the Gulf Stream due to atmospheric influences.

The periods during which changes in sea surface temperature might impact the medium-range atmospheric predictions are during the autumn and the spring transition in the ocean. During these periods, there are warm, shallow ocean mixed layers that may deepen and cool rapidly in response to atmospheric storms (Camp and Elsberry, 1978; Elsberry and Camp, 1978; Elsberry and Raney, 1978; Elsberry and Garwood, 1978). During the winter, the ocean mixed layer is so deep that the effect of even major storms on the sea surface temperature is relatively minor. Cyclones occurring during the summer tend to be less intense, and thus elicit a relatively smaller response even though a warm and shallow ocean mixed layer exists.

Another region of large upward fluxes of heat and moisture from the ocean is near the polar ice margins. These fluxes rapidly modify the lower tropospheric flow off the ice-covered region. Carleton (1981) reviews the satellite-observed cyclogenesis events in the southern hemisphere. Cyclones with relatively small horizontal scales frequently occur in the basic current downstream from the ice edge. Many observational and numerical studies have discussed the development of intermediate scale cyclones within polar air masses (e.g. polar lows). Because the cyclones develop within the cold air mass, the modifications in the air stream due to air-sea interaction must be important. However, when the ocean mixed layer is relatively deep, the corresponding modifications in the sea surface temperature will not be important.

In summary, the atmospheric response to the ocean is regional and seasonal in nature. Equatorial regions, western boundary regimes, ice margin zones and the mid-latitude regions during the autumn and spring are sensitive areas. As indicated in Exhibit 1, the ocean effects will not be felt immediately. These secondary effects assume more importance as the atmospheric predictions are extended to 10 days and beyond. However, the atmospheric predictability is also not high in these time ranges. The question is then what is to be gained (or lost) by including a time-dependent representation of the sea surface temperature in a coupled atmosphere-ocean model.

2.3.2 Nature of the Coupling. Numerical atmospheric prediction models generally only require a specification of the sea surface temperature distribution with time. The planetary boundary layer (PBL) fluxes of heat, moisture and momentum are all affected directly or indirectly by the specification of the sea surface temperature. Because of the close coupling of the PBL and the Arakawa-Schubert cumulus parameterization scheme in the NOGAPS model, the surface heat and moisture fluxes are felt through the lower tropospheric layers on time scales of less than a day.

As indicated above, the atmospheric forcing required for the ocean prediction models includes the surface stress vector, surface heat and moisture fluxes, precipitation rates and short wave and long wave radiative fluxes. The simplest one-dimensional, mixed layer models require only the magnitude of the surface stress. The TOPS model requires the direction of the stress as well. Inclusion of oceanic advection processes in any of the models requires knowledge of the stress vector. Just as the surface fluxes of heat and moisture rapidly affect the atmospheric boundary layer, these fluxes have a direct effect on the upper layer of the ocean. Periods of upward net heat flux from the ocean (at night, and especially during periods of high winds, and generally during the cooling phase of the seasonal cycle) are associated with deepening mixed layers in the ocean. Conversely, periods with net downward heat flux (during the day, and especially intervals between storms, and during the heating phase of the seasonal cycle) tend to be associated with shallow ocean mixed layers. When the heating rate is large and the wind-induced mixing is small, the sea surface temperature can increase by 0.5-1.0°C in less than a day. However, these high temperatures are contained in a near-surface layer that is only a few meters in thickness. In nature, the heat contained in these layers is normally transferred to the atmosphere, or to greater depths in the ocean, during the night. Thus, the high sea surface temperatures predicted by TOPS during the afternoon should not be regarded as representative of daily values.

It should also be emphasized that the distribution of cloudiness in regions of light surface winds is an important factor in the predicted sea surface temperature increase. Areas of minimum cloudiness will have greater downward solar flux to the ocean and larger increases in sea surface temperature during the day. The upward heat flux will dominate at night, thereby offsetting the gain during the day. Thus, to predict correctly the diurnal changes in sea surface temperature in regions of light surface winds, one must be able to predict the amount and type of clouds including any diurnal variations. No extensive and detailed verifications of cloud predictions from the NOGAPS model have been made.

Ocean models that include a salinity prediction require specification of the evaporation and precipitation rates. Formation of shallow ocean mixed layers occurs (Price, 1979) in regions where the precipitation greatly exceeds the evaporation. Since this generally occurs in storms where the wind speeds are increased, and the cloud cover is a maximum so that the net surface heat flux is upward, these shallow layers do not persist. One would not expect much of an effect on the thermal structure prediction, except perhaps in the Intertropical Convergence Zone where the precipitation exceeds the evaporation in a region of light surface winds and potentially large downward solar flux to the ocean. In the middle latitudes the excess of precipitation over evaporation leads to persistent introduction of fresh water at the surface of the ocean. This inhibits deepening of the oceanic mixed layer. No evaluations have been made regarding the accuracy of NOGAPS precipitation estimates over the mid-latitude oceans. Nor have any studies been made of the ability of the TOPS model to treat the salinity effects with either real or predicted precipitation rates.

In summary, the NOGAPS model requires only sea surface temperatures as a function of time. The surface heat, moisture and momentum fluxes, the radiative fluxes and the precipitation rates that are calculated by the NOGAPS model are required to drive the ocean model. The response of the ocean to this atmospheric forcing may have a strong diurnal period, as well as a synoptic and seasonal time scale. The amplitude of the diurnal response in sea surface temperature is critically related to the ability of the atmospheric model to predict cloud amounts and types in regions of light surface winds. These comments apply throughout the tropical regions and into the subtropics and mid-latitudes during the heating season (generally from April to September). The seasonal increase in sea surface temperature is intermittent - alternating periods of cooling during stormy periods and warming between storms (Elsberry and Raney, 1978). An accurate coupled atmosphere-ocean model for time scales of 10 days would have to be able to treat these feedback processes on diurnal and synoptic time scales.

<u>2.3.3 Mechanics of Coupling Atmosphere-Ocean Models</u>. The time step in the NOGAPS model is of the order of 5 minutes. However, the heating package is only called each 30 minutes, and it is on this time scale that the sea surface temperature affects the atmospheric prediction. The time step in the ocean models is approximately one hour. In a fully synchronous integration of the atmospheric and ocean models (case C below), all of the heat, moisture, momentum, radiation and precipitation rates

Case A. Minimal Feedback from Ocean to Atmosphere

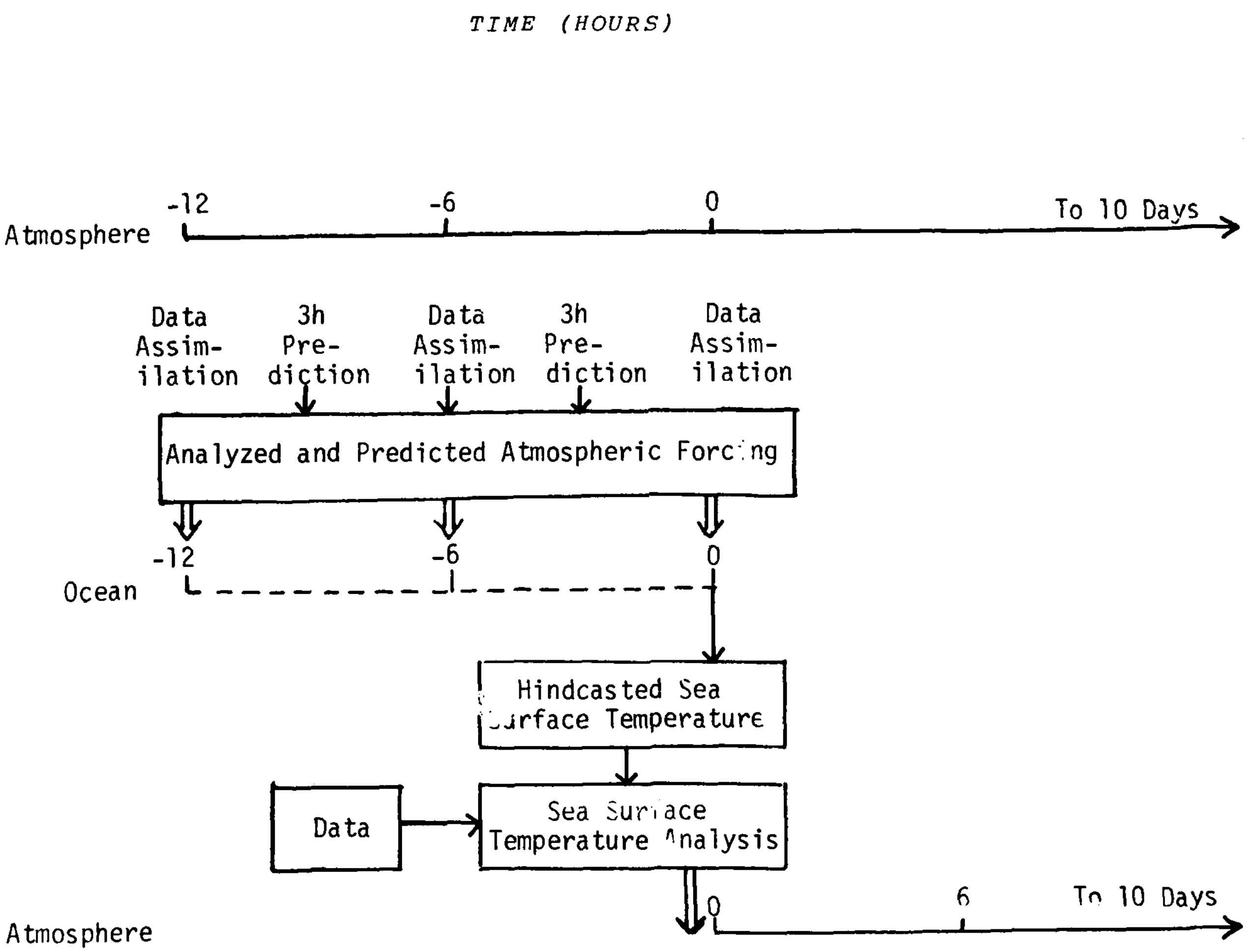
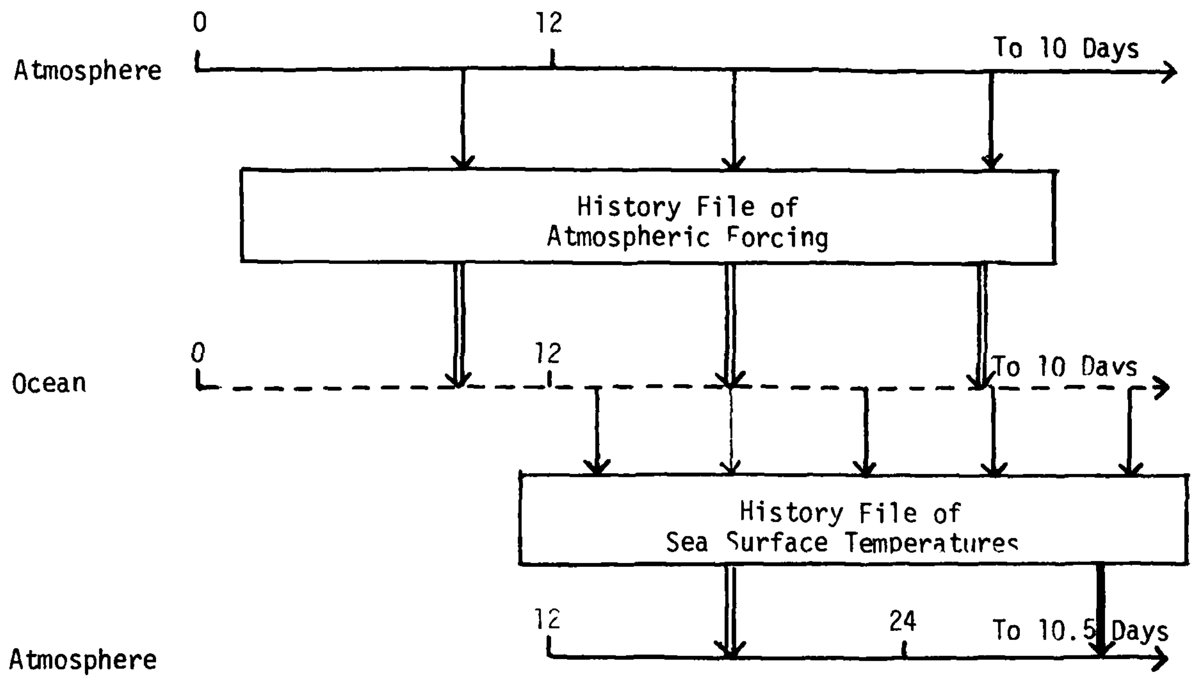


EXHIBIT 3. Possible types of ocean interaction with a medium-range atmospheric prediction model.

Case B. Non-Synchronous Coupling



Case C. Synchronous Coupling

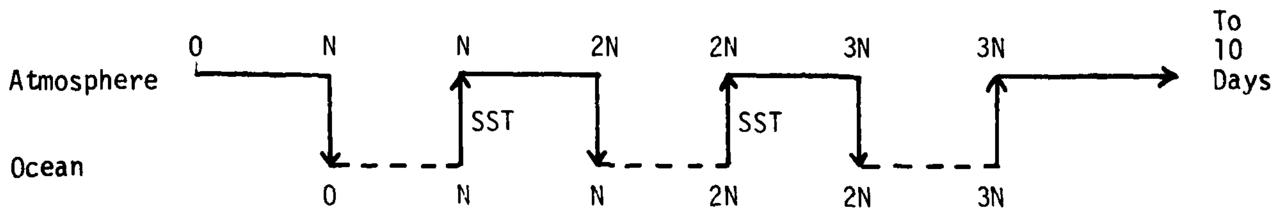


EXHIBIT 3. Possible types of ocean interaction with a medium-range atmospheric prediction model (continued).

would be provided each time step of the ocean model. The TOPS model has not been integrated with such a detailed time history of the atmospheric forcing, because of the expense of outputting and archiving these NOGAPS fields (actually only the Northern Hemisphere Primitive Equation Model has been used with TOPS). At best, only the instantaneous values of the fluxes at 6 or 12 hour intervals (except for precipitation which is accumulated over 12 hours) have been available for the ocean model. Thus some form of smoothing and interpolation to the one-hour time step of the ocean model has been used.

Some possible types of interaction between the ocean and the atmospheric prediction models are illustrated in Exhibit 3. Case A is expected to be similar to the operating procedure when both of the NOGAPS and TOPS models are accepted. The history file of the NOGAPS atmospheric forcing for the first 12 hours (a more likely interval will be 24 hours to account for diurnal variations) prior to the initial time will be used to drive the TOPS model in a "hindcast" mode. To obtain the best possible atmospheric forcing for this TOPS short-term hindcast, it is recommended that the forcing at t=0, 6 and 12 hours be obtained from a series of "analyses" derived as a part of the four-dimensional data assimilation cycle for the atmospheric model. Because the atmospheric forcing must be specified at least each three hours, the intermediate three hourly fluxes must be derived from short integrations of the NOGAPS model (this is a part of the data assimilation cycle). The primary objective of the ocean model as proposed here will be to provide the best possible first guess of the sea surface temperature for the next analysis/prediction cycle of the atmosphere model. Of course, the ocean model may also be useful as a prediction of the subsurface thermal structure. However, that aspect is not the concern of this review. As indicated in the sketch, all available conventional and satellite observations will be combined with the model hindcast to form the sea surface temperature analysis at t=0. Finally, a key aspect of this case is that the sea surface temperature will be held fixed throughout the subsequent 10-day atmospheric prediction. Thus the "minimal feedback" from the ocean model to the NOGAPS model will be via the first guess to the sea surface temperature analysis.

Case B in Exhibit 3 has two key aspects. First, it presumes that a sophisticated ocean model will be developed and run for its own sake. When such a useful and accurate ocean model becomes available, the limiting feature will be the length of time that accurate atmospheric forcing is available. Thus, it is assumed that the ocean model will be extended to 10 days (or whatever interval the NOGAPS model predictions are available). In this case, the ocean model will be run separately after the atmospheric model has created a history file of the required forcing. The second aspect is that the time-dependent sea surface temperatures will be specified during the next atmospheric prediction cycle. The strategy in this "non-synchronous coupling" of the models is to provide a time-dependent sea surface temperature distribution, without impeding the progress of the atmospheric model integration. Although this temperature distribution would be based on the prior atmospheric model cycle, it might be expected to include the major tendencies associated with the diurnal and synoptic (storm) forcing.

A "synchronous coupling" of the ocean model with the atmospheric model is illustrated schematically in Case C in Exhibit 3. The atmospheric model is first integrated N time steps. The history of the atmospheric forcing during this interval is used to integrate the ocean model to the same time level. The newly-derived sea surface temperature then becomes the lower boundary condition for the atmospheric model integration. This strategy requires an essentially synchronous integration of the two models. The coupling may be considered <u>fully</u> synchronous if the time interval N is equal to the time step of the ocean model (which is larger than for the atmospheric model). However, other intervals may be considered to reduce the number of times the two models would be interchanged in the computer. A good choice for N might be 24 hours, because this would provide a suitable interval for averaging the sea surface temperature (to eliminate the diurnal, smallest and least predictable scales) that is to be provided during the next day of the atmospheric model integration. For these longer intervals, the coupling is not really complete, because the sea surface temperature would be held fixed during the (say) 24-hour periods of the atmospheric integration. This aspect is somewhat similar to Case A, although the parallel evolution of the atmospheric and oceanic states is similar to Case B.

2.3.4 Synchronous Coupled Atmospheric-Ocean Models. Examples of the fully synchronous coupling (as defined in Exhibit 3) of atmospheric models with an active ocean model include the hurricane-ocean model of Chang and Anthes (1979) and the sea breeze model of Clancy (1977). The differential surface temperature response between land and the adjacent ocean to diurnal heating was treated by Clancy (1977). Such local scale effects are well below the resolution of the NOGAPS model and will not be examined further. The hurricane-ocean interaction provides a good model test because of strong response elicited in the ocean by the intense, small scale wind field. Ocean mixed layer models (even without advection) predict the first-order response in sea surface temperature, although more complex ocean models (Price, 1981; Adamec *et al*., 1981) are required to describe the interaction between mixing and advection processes. The horizontal scale of the hurricane and its associated wake in the ocean are too small to be predicted by the NOGAPS/TOPS models.

The closest prototype experiment to a coupled NOGAPS-TOPS model is the work of Wells (1979). Because of its importance, this attempt at coupling will be described in some detail.

The Australian spectral (9 levels, resolution to wave number 15) Southern Hemisphere general circulation model was used for a January (summer) situation. A key assumption is that the cloud distribution is fixed at climatological January conditions. The ocean mixed layer model is based on the partially penetrative slab model of Gill and Turner (1976). Special features include the dissipation of turbulent kinetic energy with depth, horizontal advection of temperature and salinity in the mixed layer, and uniform vertical advection below the mixed layer to the bottom of the model (200 m). The ocean advection was based on an Ekman balance with the imposed wind stress from the atmospheric model. In this regard the model is similar to the advective version of the TOPS model, except that the latter also includes an option for a climatological geostrophic current.

The coupling of the models is identical to Case C of Exhibit 3 with N being 4 days (additional runs with N=1 day gave similar results averaged over one month). During the atmospheric model integration the sea surface temperature is fixed. The history file of surface fluxes (calculated each hour) are averaged over the 4-day period before being used to drive the ocean model. This averaging eliminates most of synoptic (cyclone) scale variability (no diurnal cycle was included in this experiment). This coupling procedure and the turbulence closure scheme in the ocean model assures a slowly evolving ocean temperature distribution. All interpretations of the ocean results were based on time averages over 32 days after an adjustment period of 40 days. The adjustment period avoids the difficulty of initialization of the Southern Hemisphere ocean thermal structure.

Some care must be exercised in applying the results of the Wells coupled model to operational prediction. Some potentially significant and relevant results include:

(1) The planetary boundary layer formulation in the

atmospheric model is an important factor (Chang and Anthes, 1979, obtained a similar result in their hurricane-ocean experiment).

(2) Provision for a variable oceanic mixed layer depth is necessary for distributing properly the heat input (recall that this is a mid-summer simulation, and the ocean is warming at all latitudes).

(3) Equatorial and polar ice zones are regions in which large differences occur in the ocean model with and without advection. (The wall boundary condition with the Ekman-type currents may have contributed to the large equatorial differences, whereas the fixed cloud and ice edge could account for the polar differences.)

(4) The coupled model achieves a "climatology" of surface winds which are rather poor from the Equator to 45°S. The sea surface temperatures are sensitive to areas of light winds.

Although the Wells experiment does not include diurnal and synoptic-scale variability, an initialization, or a fully synchronous integration, the demonstration of stable, extended (72 day) integration is an important milestone. Similar experiments should be done with winter conditions of strong atmospheric forcing and deep ocean mixed layers.

2.3.5 Results from Coupled General Circulation - Ocean Models. The advanced atmospheric general circulation models (such as at UCLA, GFDL, etc.) and the operational model at the European Center for Medium-Range Weather Forecasts (ECMWF) provide guidance as to benefits to be expected from improved atmospheric models. There are no comparable research versions of coupled models to use as a standard for assessing the benefits to be obtained from an operational coupled model. The research emphasis in most large-scale coupled models has been on annual or longer time scales. Climate sensitivity studies, and particularly the response of the atmosphere to continued increases in carbon dioxide, require a coupled atmosphere-ocean model. Active research with coupled models is in progress at GFDL, Oregon State University (OSU), NCAR and other centers. Even though the emphasis in these research versions is on climate time scales, some indications can be gained regarding the required horizontal and vertical resolution, and what physical processes must be included for realistic results. The presumption is that any operational model (coupled or not) will tend to a particular "climate" as the model "forgets" the initial state, and that this "climate" should be near the observed climate. This may be regarded as a necessary, but not sufficient, condition for a forecast model.

Some cautionary statements are appropriate. Coupled atmosphere-ocean models are complex and expensive, especially when the integrations are intended for climate studies. Consequently, the atmospheric models generally have coarse horizontal and vertical resolution. These models have well known limitations regarding ability to resolve small scale cyclones and their associated vertical fluxes. A common characteristic of the atmosphere models is the specification of a fixed cloudiness distribution (normally a zonal mean). These atmospheric models will not contain the diurnal and (complete) synoptic forcing that will be important in medium-range forecasts. As we will indicate below, the ocean models are also rather simplified.

The GFDL group has the longest history of coupled atmosphere-ocean studies, beginning with Manabe _et al_. (1975) and Bryan _et al_. (1975). More recent studies include those of Manabe, Bryan and Spellman (1979), Manabe and Wetherald (1980) and Manabe and Stouffer (1980). One of the interesting aspects of the early papers is the asynchronous mode of integration of the atmospheric and ocean models. For example, in Manabe _et al_. (1979), the ocean model is advanced many years using one month of atmospheric forcing, so that a 4.2 year integration of the atmospheric model corresponds to 1200 years in the ocean. A set of annual means of the monthly sea surface temperatures is calculated to feed back to the atmospheric model. A computational instability was discovered that is associated with the heat exchange calculation that is made in the atmospheric model for use during the ocean cycle. Calculating separately the heat flux in both models (monthly mean atmospheric values used in the oceanic portion and monthly mean temperature in the atmospheric part) eliminated the instability. The Manabe and Stouffer version of the ocean model includes a simple mixed layer to avoid some of the difficulties with unrealistic layer depths in the Manabe et al. (1979) version. The approach in the later model is to establish an effective (uniform) depth of the seasonal thermocline. Although one could calculate a latitudinal variation of this depth, it was set at 68 m throughout the ocean. Other studies (e.g. Thompson, 1976) have used a single value for the depth over which the atmospheric forcing is distributed. This generally results in a reduction in the amplitude of the seasonal variation in sea surface temperature. Other features of this study are the delayed feedback to the atmosphere of the heat stored in an anomalously deep ocean layer, and the complex interaction with the sea-ice boundary in the polar regions. Both of these interactions with the atmosphere suggest that errors can be accumulated in the ocean that might later produce erroneous atmospheric circulations if the incorrect ocean predictions are not updated with adequate observations.

An effort at GFDL which is closer to the topic of this review is the research by Miyakoda and his group toward producing 30-day atmospheric forecasts. An attempt is being made to adapt K. Bryan's ocean model for use with the atmospheric models. An increase in vertical resolution from 25 m to 2.5 m near the surface produced more realistic sea surface temperature fields and a more realistic vertical temperature distribution (Miyakoda, personal communication). They are also concerned that the ocean be in balance with the atmospheric forcing. Their present scheme is to integrate the ocean model with annually varying atmospheric forcing for seven cycles. They then run the atmospheric model with the new monthly sea surface temperature fields, and repeat the cycle. The goal is to have a stable ocean "climate" on which to superpose sea surface temperature anomalies, and then determine their effect on the 30-day atmospheric forecasts.

The present status of the Oregon State University coupled model is summarized in Schlesinger and Gates (1981), which also lists earlier versions of the ocean model (Kim, 1976, 1979; Kim and Gates, 1980; Heald and Kim, 1979; Pollard, et al., 1980). The present version contains a (fixed depth - 60 m) mixed layer and a sea ice model. This coupled model is fully synchronous with a one hour time step. As was the case with the fixed depth model of Manabe and Stouffer (1980), the sea surface temperatures are too low in the summer and too high in the winter. One exception is near the ice margin during the southern hemisphere summer where the temperature is too high. The ocean model did not perform as well when it was coupled to the atmospheric general circulation model as it did when it was driven by prescribed (and non-interactive) atmospheric conditions. There appears to be a positive feedback loop of errors in the coupled version. They also conclude that on monthly time scales: "the simulated surface air temperatures and precipitation rate clearly demonstrate the important control exerted by the sea surface temperature on the atmospheric climate, at least over the ocean".

2.3.6 Additional Research Efforts. The British Meteorological Office has been developing a coupled atmosphere-mixed layer ocean model that includes sea ice (Bottemley and Gordon, 1980). Only non-interactive tests of the ocean model have been completed as of June 1981. The same sea ice distribution is applied in the ocean model as is used in the atmospheric general circulation model from which the history file of forcing is derived. Regions in which the simulated wind speeds are low for extended periods (more than a week) produced extremely large (30-40°C) sea surface temperature increases. Later experiments are planned to include a fully interactive atmosphere-ocean model with a predicted sea ice distribution.

Two research efforts that do not involve coupled ocean models will also be briefly discussed. These efforts are interesting partly because of the atmospheric models used to generate the forcing. In the first case (Arpe, 1981), the ECMWF model is used and, in the second (Sandgathe, Elsberry and Winninghoff, 1982), the NOGAPS model is used. One of the curious aspects of the ECMWF model is that a monthly, climatological sea surface temperature is used as a lower boundary condition. Arpe (1981) describes some experiments in which daily sea surface temperatures (still fixed in time) are used. The ECMWF method of verification, which extends over the entire hemisphere north of 20°N and throughout the troposphere, tends to obscure the effects due to a different sea surface temperature. Examination of the lower troposphere fields (where one would expect to see the effects) over the ocean shows changes after 3-4 days compared to the model with monthly mean sea surface temperatures. The stronger cyclogenesis and faster movement are in the same sense as Sandgathe (1981) found in the NOGAPS model when the surface fluxes of heat and moisture were removed. This indicates that the ECMWF model with excellent horizontal and vertical resolution and with sophisticated parameterizations does respond to discrete (and fixed in time) changes in sea surface temperature of only 1-3°C.

Sandgathe _et al_. (1982) have used the history file from the NOGAPS model initialized from idealized conditions to drive the one-dimensional mixed layer model of Garwood (1977). No salinity effects and no advection is included in the ocean model. The simulation corresponds to autumn conditions of warm and shallow mixed layers in the northern hemisphere ocean, and cold and deep layers typical of late winter in the southern hemisphere. In the northern mid-latitudes there is a decrease in sea surface temperature of 1-2°C in the wake of the extratropical cyclones. Even though the southern hemisphere cyclones are more intense, the response in the ocean is much less because of the initial depth of the mixed layer. These results are consistent with earlier observational and numerical studies (Camp and Elsberry, 1978; Elsberry and Camp, 1978; Elsberry and Raney, 1978).

The surprising result in these preliminary tests is the ocean response in the equatorial regions. This is an area of net downward heat flux over the period of a day, and there are light wind speeds in this simulation. As expected, the heat is accumulated in a shallow layer with increases in the zonal mean sea surface temperature of more than 2°C in 5 days. This type of warm equatorial anomaly would be expected to induce extremely vigorous deep convection if the new sea surface temperatures were fed back to the NOGAPS model. A second surprising aspect is the amount of east-west variability that appears in the predicted sea surface temperature. Harmonic analysis indicates that there is an east-west variance of approximately 0.3°C in high wave numbers (18 through 30) in the equatorial region, whereas there is little real variance in these wave numbers in the mid-latitudes. This type of variability must be due to the dependence on cloudiness of the downward solar flux in the NOGAPS radiation package. Rather than fixed (usually zonal mean) cloudiness that is characteristic of the studies described above, there is an additional degree of freedom introduced into the ocean model by the predicted cloudiness-radiative flux relation. One might expect a highly complex interaction if these effects are included in a fully coupled ocean model. Oceanic regions that are relatively cloud-free (and with light surface winds) would have increasing sea surface temperatures relative to adjacent cloudy regions. If this produces temperature increases of the magnitude indicated in the non-interactive experiment, these warm anomalies would destabilize the troposphere and deep convection would result. This would subsequently diminish the downward heat flux to the ocean. There is little doubt that this type of negative feedback occurs in nature, but it is uncertain what the magnitude or intensity of the feedback is. The point here is that an incorrect sea surface temperature prediction of perhaps 1°C could occur due to incorrect atmospheric forcing (due to cloudiness effects or wind stress). In the tropics, the effect of this erroneous surface temperature would be spread vertically throughout the troposphere, and perhaps into the mid-latitudes in certain situations. Thus, the omission of this negative feedback in a one way interactive model in which the atmospheric predictions influence the SST analyses (and not the forecasts) could have serious implications. One can see the possibility of unrealistic air-sea interactions in regions with warm (or cold) SST anomalies that are held constant during the entire period of the atmospheric forecast. Considerable research is necessary on the proper representation of the diabatic and frictional processes in coupled models before one can be confident of a beneficial effect on the atmospheric predictions.

Although the Sandgathe et al. (1982) experiments must be regarded as very preliminary, they indicate that using the most advanced atmospheric models can introduce additional degrees of freedom into coupled models. Small scale features induced in the ocean surface layers due to cloud feedback processes may have both positive and negative benefits. There is definitely less predictability on these space and time scales, and it is perhaps as likely that this scale of atmospheric-ocean feedback will be modelled incorrectly as correctly. Until these effects are better understood, especially with regard to any deleterious effects on the atmospheric prediction, one must advise caution in proceeding to fully coupled atmosphere-ocean models.

2.4 Stratospheric Forecast Modeling.

2.4.1 Observed Stratospheric Circulation. The mean stratospheric circulation has strong westerlies in the winter hemisphere and strong easterlies in the summer hemisphere. The basic flow is primarily driven by differential heating due to the absorption of solar ultraviolet energy in the ozone layer centered at about 50 km ($\sim$1 mb) and infrared emission to space due to ozone, carbon dioxide and water vapor. This heating variation drives a mean meridional circulation with rising motion over the summer pole, where there is heating, and sinking over the winter pole, where there is cooling. The direct circulation from summer pole to the winter pole intensifies the easterlies in the summer hemisphere and the westerlies in the winter hemisphere through the coriolis torque. Since the total circulation is in the process of reversing at the equinoxes, the winds are much weaker than at the solstices. However, the seasonal reversal is not a simple annual cycle and, in fact, the semi-annual oscillation is dominant near the equator.

In the summer hemisphere all disturbances in the extratropical stratosphere damp rapidly with height. However, quasi-stationary disturbances composed of wave numbers 1 and 2 are found in the winter stratosphere. These disturbances vary with periods of the order of two weeks, and they are sometimes connected to the major stratospheric warmings. During the latter events, the westerly winds at high latitudes are eliminated over one to two weeks and the temperatures at the pole rise rapidly. Although the total breakdown of the polar night vortex occurs only every other year or so, minor warmings are much more common. Quirox (1977) has presented an example of large tropospheric changes associated with a sudden stratospheric warming event.

The tropical stratosphere contains mixed Rossby gravity waves and Kelvin waves which play an important role in the quasi-biennial oscillation [Holton, 1975]. However, because of the long period of the oscillation, and the small amplitude of the waves in the troposphere, it is unlikely that these waves need to be modeled for medium range forecasting. Other more important tropical tropospheric prediction problems need to be addressed prior to treating the tropical stratospheric waves. Thus, this portion of the review will be directed toward understanding and modeling troposphere-stratosphere interaction in the extratropics during winter.

2.4.2 Extratropical Stratospheric Dynamics. Nakamura (1976) and Kirkwood and Derome (1977), used simple linear models to investigate the influence of the upper boundary condition on topographically forced long waves. They found very large errors in the troposphere when the mean wind conditions produced vertically propagating waves, and when the stratosphere was poorly resolved. Thus it is important to review the theory of stratospheric wave propagation and mean flow interaction in order to better understand the modeling problems.

Many dynamical effects can be explained following Holton (1980) and Holton and Dunkerton (1978) with the linearized quasi-geostrophic potential vorticity equation:

$$\left(\frac{\partial}{\partial t} + \bar{u}\frac{\partial}{\partial x}\right) q' + v' \frac{\partial \bar{q}}{\partial y} = -S', \tag{1}$$

where

$$q' = \frac{1}{f}\nabla^2 p' + \frac{f}{N^2}\frac{\partial^2 p'}{\partial z^2} , \tag{2}$$

$$\frac{\partial \bar{q}}{\partial y} = \beta - \frac{\partial^2 \bar{u}}{\partial y^2} - \frac{f^2}{N^2}\frac{\partial^2 \bar{u}}{\partial z^2} , \tag{3}$$

$$S' = \frac{\partial F_y{}'}{\partial x} - \frac{\partial F_x{}'}{\partial y} - \frac{f}{N^2}\frac{\partial Q'}{\partial z} . \tag{4}$$

Here N is the Brunt-Vaisala frequency, Q is the heating and F_x and F_y are frictional components. The Boussinesq approximation has been used, and the bar and prime indicate x-mean and perturbation quantities, respectively. The mean potential vorticity equation can be written:

$$\frac{\partial \bar{q}}{\partial t} = -\frac{\partial}{\partial y}\left(\overline{q'v'}\right) - \bar{S} , \tag{5}$$

where

$$\bar{q} = \frac{1}{f}\frac{\partial^2 \bar{p}}{\partial y^2} + \frac{f}{N^2}\frac{\partial^2 \bar{p}}{\partial z^2} + f , \tag{6}$$

$$\overline{q'v'} = -\frac{\partial}{\partial y}(\overline{u'v'}) + \frac{f}{N^2}\frac{\partial}{\partial z}(\overline{v'\theta'}) \quad . \tag{7}$$

Vertical wave propagation can be easily investigated by neglecting S' and setting $\partial p'/\partial y = \partial \overline{u}/\partial y = 0$. Then wave solutions of the form

$$p' = P_0\ (z) \exp\ [ik(x-ct)] \quad , \tag{8}$$

can be found by substitution into (1). The resulting vertical structure equation is

$$\frac{d^2P_0}{dz^2} + m^2P_0 = 0 \quad , \tag{9}$$

where

$$m^2 = [\beta/(\overline{u} - c) - k^2]\ N^2/f^2 \quad . \tag{10}$$

There will be vertical propagation if $m^2 > 0$ which gives the condition

$$0 < \overline{u} - c < \beta/k^2 \quad , \tag{11}$$

which is a slightly simplified version of the relation derived by Charney and Drazin (1961). This shows that stationary waves (c=0) can propagate vertically only when the mean wind is westerly and the wavelength is long enough so that $\beta/k^2 > \overline{u}$. This explains why stationary disturbances are only found in the winter stratosphere ($\overline{u} > 0$) and why they are of large scale. Equation (10) also shows that waves cannot propagate past a critical level where $\overline{u} = c$.

An equation for $\overline{q'v'}$ can be obtained by multiplying (1) by q' and averaging with respect to x:

$$\overline{q'v'} = -\left[\frac{1}{2}\frac{\partial}{\partial t}(\overline{q'^2}) + \overline{S'q'}\right] / \partial\overline{q}/\partial y \quad . \tag{12}$$

If $\partial\overline{q}/\partial y \neq 0$, there will be potential vorticity transport by the eddies only if the eddies are changing in amplitude or they are subject to dissipation. For steady adiabatic waves, $\overline{q'v'} = 0$, and consequently there is no net forcing of the mean flow by the eddies [see (5)]. This is the Charney-Drazin non-acceleration theorem [Charney and Drazin (1961)].

2.4.3 Steady Stratospheric Waves. Matsuno (1970) computed steady stratospheric waves which were forced from below. He used the linearized quasi-geostrophic equations with a realistic mean flow and monthly mean 500 mb heights for the lower boundary condition. Newtonian cooling and Rayleigh friction terms were included with small coefficients in order to avoid singularities at critical levels. The solution for wave number 1 compared reasonably well with observations, but wave number 2 was too weak in the upper levels. The solutions tilted westward and equatorward with height.

Both the computed and the observed solutions showed propagation generally along the axis of maximum wind in the latitude-height plane. This behavior was not expected from the theory of Charney and Drazin (1961), which predicted a maximum vertical penetration during the equinoxes when the westerlies are weaker. In fact, Dickinson (1969) has found that stationary waves during the equinoxes are rapidly damped with height by a reasonable Newtonian cooling. Simmons (1974a) in an analytic study showed that vertical propagation in strong westerlies is enhanced when the latitudinal scale of the mean flow and that of the disturbance are similar. In particular, if the amplitude P_0 in (8) is a function of y and z, the following terms

$$\overline{u}\,\frac{\partial^2 P_0}{\partial y^2} - P_0\,\frac{\partial^2 \overline{u}}{\partial y^2} \quad ,$$

which arise from substitution of (8) into (1), will tend to cancel, which reduces the effect of the mean advection. Dickinson (1969) and Simmons (1974a) have found that when strong westerlies are present, the vertical penetration of the waves is actually increased by Newtonian cooling.

Schoeberl and Geller (1977) carried out a more extensive investigation of steady stratospheric waves with a variety of conditions. With realistic mean winds, the solutions fit the observations well in the lower stratosphere. In general, it was found that the wave structure was very sensitive to the winds in the polar night jet.

2.4.4 Numerical Modeling of Sudden Stratospheric Warmings. The observational structure and dynamical mechanisms of sudden stratospheric warming have been reviewed by Holton (1975, 1980) and Schoeberl (1978). This phenomena is an example of strong interaction between the troposphere and the stratosphere which is often associated with tropospheric blocking. Quirox (1977) has presented an example of this interaction. Dopplick (1971) obtained energy variations with an approximate period of two weeks, and Koermer and Kao (1980) found long wave oscillations in wave numbers 1 and 2 of 10-20 days in the lower stratosphere during winter. It appears that proper modeling of the sudden stratospheric warmings and the other winter stratospheric oscillations is necessary for improved forecasting of troposphere-stratosphere interaction. This subsection will review various models which have been applied to the sudden stratospheric warmings. These models use specified boundary conditions at the tropopause.

Earlier explanations for the sudden warmings concentrated on baroclinic or barotropic instability of the polar night jet, which contains large vertical and horizontal wind shears [Murray (1960), McIntyre (1972), Simmons (1974b)]. These studies show that these instabilities are possible, but they have the wrong horizontal or vertical scales for the observed structure of the warmings.

Matsuno (1971) formulated the first model that explained the main features of the sudden stratospheric warmings. He employed a time dependent forcing at the tropopause which forced an upward propagating, transient wave. The Charney-Drazin non-acceleration theorem [see (5) and (12)] states that adiabatic waves must have changing amplitude to affect the mean flow. For example, adiabatic steady waves which satisfy (1), and which are forced from below, transport heat to the north (i.e. $\overline{v'\theta'} > 0$); but according to this theorem, they cannot affect the mean potential vorticity. In this case, the heat flux convergence is exactly cancelled by the adiabatic cooling from the mean meridional circulation. For the general case, the mean meridional circulation operates to maintain geostrophic balance in the zonal flow, and this circulation is determined by the distribution of $\overline{v'\theta'}$, $\overline{u'v'}$ and $\overline{Q}$.

Matsuno forced a wave by rapidly increasing the height at the lowest level in the model ($z \sim 10$ km) by 300 m and holding it fixed thereafter. The disturbance was a maximum at 60° latitude and it decreased to zero at the pole and at 30°. Matsuno proposed the following sequence of events. As the energy propagated upward and a northward heat flux developed, the mean temperature increased in the polar region. The mean meridional circulation developed with rising motion in the polar regions and a southward mean me ional flow. The adiabatic cooling associated with the rising motion partially offset the polar warming caused by $\overline{v'\theta'}$, and the meridional flow decreased the westerly mean wind through the coriolis torque. This process increased with height because of the smaller densities aloft. Matsuno proposed that when the westerlies were reduced to zero at a certain level, the critical level effects greatly accelerate this process. Since the vertical energy propagation is blocked at a level where $\overline{u} - c = 0$ (in this case $c=0$), the heat flux $\overline{v'\theta'}$ would be zero above that level. In order to maintain geostrophic balance in the mean zonal wind, a very strong southward mean flow is required at that level. Through the coriolis torque, the mean flow is then rapidly decelerated at that level and this process caused the $\overline{u} = 0$ line to descend rapidly, thereby speeding the breakdown of the vortex and accelerating the warming process.

The numerical solutions with wave numbers 1 and 2 showed a wave growth phase of about 10 days, after which the interaction with the mean flow began to decrease the mean wind in the polar night jet, and increase the temperatures in the polar region. After about 20 days, easterly mean winds appeared, and the polar night vortex was replaced by a high pressure area. The maximum temperature increases were 40°C and 80°C for wave numbers 1 and 2, respectively. These numerical solutions are quite similar to observed sudden stratospheric warmings.

Holton (1976) repeated Matsuno's experiments with a non-geostrophic model and obtained very similar solutions. He found that the momentum flux $\overline{u'v'}$ was important during vortex breakdown, which was also shown in an observational study by O'Neill and Taylor (1979). The Matsuno heuristic model gives a general understanding of the sudden stratospheric warming, but the actual phenomena is considerably more complex. In particular, momentum fluxes are important as well as north-south wave propagation.

Schoeberl and Strobel (1980) studied sudden stratospheric warmings with a quasi-geostrophic model which is very similar to the one used by Matsuno (1971). However, the initial mean winds were derived from steady state solutions to the basic equations which included mean heating [see Schoeberl and Strobel (1978)]. They also used more realistic heating. In general, they found that the sudden warmings are critically dependent on the vertical transmission of planetary waves, which depend on the following factors: (1) the strength of the westerlies in the lower stratosphere; and (2) the magnitude of the wave damping in that region. Major warmings occurred when the westerly winds in the lower stratosphere were strong, but not so strong that the waves were trapped at low altitudes. The magnitude of the damping determined the maximum temperature change in the warming and the total development time. However, the wave number 1 and wave number 2 warmings were quite different. In the former case a critical line ($\bar{u}$=0) formed in the polar region advanced southward, while in the latter case, the critical line developed first in the equatorial region and advanced northward.

Lordi, Kasahara and Kao (1980) investigated the sudden stratospheric warmings using a spectral primitive equation model, which is otherwise generally similar to the models developed by Matsuno (1971), Holton (1976) and Schoberl and Strobel (1978), except that direct interaction is allowed between wave numbers 1 to 4. This study shows that the nonlinear interaction between wave numbers 1 and 2 are quite important during the sudden stratospheric warmings.

2.4.5 Numerical Modeling of the Mean Stratospheric Circulation. In this subsection models of the mean stratospheric circulation are reviewed. Accurate modeling of the mean circulation for a medium-range stratospheric prediction model is important because there could be a significant erroneous trend during 10 days, if the model were to produce poor mean fields. Also a good model will be required to improve analysis of the stratospheric fields through data assimilation, since upper atmosphere data are very sparse.

Leovy (1964) developed an analytic model of the annual variation of the zonal mean stratospheric and mesospheric circulation. The mean zonal wind was assumed to be in geostrophic balance. Eddy heat and momentum transports were approximated with linear damping terms. With a reasonable heating distribution, he obtained the correct basic circulation with easterlies in the summer hemisphere and westerlies in the winter hemisphere.

Schoeberl and Strobel (1978) treated the steady state circulation with nearly the same model as used by Leovy (1964), except they used the radiative heating parameterization that was developed by Strobel (1976). A series of solutions were obtained with different values of the Rayleigh frictional coefficient, k_r. These experiments show that k_r^{-1} needs to be 10 days or less to get reasonable mean winds at the solstices: otherwise the winds in the polar night jet are much too strong. Schoeberl and Strobel found the best fit with observations when the following expression is used:

$$k_r = k_0\, e^{z/c} \quad , \tag{13}$$

where $k_0^{-1} = 30$ days, $c = 4.1$, $z = \ln(p_0/p)$ and $p_0 = 100$ mb. When wave number 2 is forced with a climatological lower boundary condition, there is a negligible effect on the mean fields in this model. With wave number 1 forcing, no steady state can be obtained due to the formation of easterlies (and thus critical lines). Schoeberl and Strobel suggest the critical layer vacillation behavior discussed by Holton and Mass (1976) prevents the steady-state solution.

Holton and Wehrbein (1980) formulated a mean circulation model of the middle atmosphere based on the primitive equations of motion. Eddies are neglected and Rayleigh friction is included. They found that values of $k_r^{-1} \sim 2$-4 days above 70 km are required to fit observations in the polar night jet. They suggest turbulence associated with gravity waves and tides can explain these large values. Lindzen (1980) estimates breaking internal gravity waves may be important in this region.

2.4.6 Numerical Models with Troposphere and Stratosphere. Smagorinsky *et al*. (1965) formulated the first general circulation model with a stratosphere. It had 9 levels with 3 levels in the stratosphere. The numerical solutions did not show significant planetary wave propagation into the stratosphere. Manabe and Hunt (1968) used basically the same model, but with twice the vertical resolution. Although there was no topography or non-zonal heating, the model did produce a minor stratospheric warming. Manabe and Terpstra (1974) returned to the 9 level vertical resolution of this model and added mountains. In their experiments wave number 1 did penetrate into the stratosphere. This model, both with and without mountains, produces polar night jet winds which were twice as large as the observed winds.

Miyakoda, Strickler and Hembree (1970) used the basic GFDL 9-level model in an attempt to predict the sudden stratospheric warming of March 1965. Forecasts made from 2 to 5 days before the vortex breakdown, predicted partial breakdown, but not the sudden warming. The lack of resolution in the stratosphere may have precluded successful treatment of the warming process.

Trenberth (1973a) developed a quasi-geostrophic spectral model which is highly truncated. This model has 9 levels extending to 70 km with a Δz in the stratosphere of about 10 km. With no topography and only zonal heating, very little energy is available in the troposphere for propagation into the stratosphere. In a second paper, Trenberth (1973b) describes a numerical experiment with topographic and thermal forcing in wave number 2. Only transient propagating waves of this scale are observed, and these cause a 20-day index cycle in the mean wind and temperature in the stratosphere. The mean zonal winds predicted in the polar night jet are much too strong, although they are somewhat smaller with wave number 2 forcing.

The NCAR model as described by Kasahara, Sasamori and Washington (1973) and Kasahara and Sasamori (1974) had 12 equally spaced levels between 1.5 and 34.5 km. The simulations without mountains produce very cold polar temperatures in the stratosphere and a polar night jet which is too strong by a factor of two. With mountains, the polar temperatures are closer to observations, but the polar night jet is almost absent.

Manabe and Mahlman (1976) studied the seasonal variation in the stratosphere with an 11 level model of which 5 levels are in the stratosphere. The mean zonal wind in the polar night vortex is too strong and the polar temperatures are too cold in the stratosphere. This model, which includes mountains, does predict stationary waves in the winter stratosphere with the proper phase. However, this model does not predict any sudden stratospheric warmings.

Schlesinger and Mintz (1979) developed a model which features ozone production and transport. The model is a 12 level version of the UCLA General Circulation Model, which uses p coordinates above 100 mb and sigma coordinates below, as described by Arakawa and Lamb (1977). The 7 levels above 100 mb are spaced according to ln p, and there is a sponge layer at the top. As with most other general circulation models, the mean winds in the polar night vortex are much too strong. Also, no sudden stratospheric warmings are observed.

O'Neill (1980) analyzed stratospheric warmings which occurred in experiments with the British Meteorological Office model, that was described by Newson (1974). It is a general circulation model with 13 levels, of which 9 are in the stratosphere. When this model is integrated with northern hemisphere winter conditions, a strong polar night jet forms which is about twice as strong as the observed jet. The temperatures in the polar regions are also far too low. After this mean flow develops, during the period from 50-78 days, the stratosphere undergoes a change which in many ways is similar to observed sudden stratospheric warmings. The breakdown process, which involves mainly wave number 1, resembles in a general way the process proposed by Matsuno (1971). However, O'Neill found that momentum fluxes are very important, but they are not considered in Matsuno's conceptual model. This is the first time sudden stratospheric warming has spontaneously occurred in a troposphere-stratosphere model.

2.4.7 Sensitivity Experiments and Forecast Comparisons. Nakamura (1976) uses a one-dimensional, quasi-geostrophic model to study the effect of stratospheric resolution and the upper boundary conditions on long waves. He uses a reasonable Newtonian cooling which damped the waves before they reached the upper boundary. He finds vertically propagating, steady state waves can be properly simulated, if the upper boundary is above 35 km (5 mb) and the vertical grid increments are as small as 1-2 km in the troposphere and 2-3 km in the stratosphere.

Kirkwood and Derome (1977) and Laprise (1978) carry out similar studies with similar results with a constant increment Δp model. However, Laprise also includes horizontal wind shear. Kirkwood and Derome (1977) point out that if a numerical prediction model produces a steady state long wave solution which is very different from the real atmosphere's solution, then a rapidly moving Rossby wave will be excited. This result is again demonstrated by Desmarais and Derome (1978) with a linearized, time-dependent model. Lambert and Merilees (1978) analyze forecast errors in a spectral numerical prediction model, and find the major contribution to short range forecast errors in the troposphere is a westward propagating Rossby wave with a 5-day period. They suggest that this Rossby wave is excited by poor stratospheric resolution. However, this wave could also be excited by inaccurate forcing of the long waves or improper tropical boundary conditions, as has been discussed by Daley, Tribbia and Williamson (1981).

All of these studies suggest that the long waves predicted by numerical prediction models should be sensitive to stratospheric resolution. The cyclone scale waves in medium-range forecasts should be affected by the incorrect prediction of long waves.

Lambert (1980) carries out a series of experiments to determine the sensitivity of an operational numerical prediction model to increased vertical resolution and the addition of stratospheric levels. The model is spectral with 29 waves, with heating and topography included. The following vertical structures are compared: (1) 5 levels, top at 100 mb; (2) 11 levels, top at 100 mb; and (3) 15 levels, top at 10 mb with resolution below 100 mb the same as model (2). Experiments are carried out for a winter case (9 January 1977) with strong stratospheric westerly winds and a summer case (11 May 1977) with stratospheric easterly winds. The differences after 48 hours between forecasts at 500 mb with the three vertical structures are sufficiently small that it is difficult to tell the maps apart. When the differences between forecasts are Fourier analyzed, the following conclusions are obtained: (1) for wave numbers less than 8, the effects of increased vertical resolution or the addition of stratospheric levels has about the same effect on the 500 mb geopotential; (2) for wave numbers greater than or equal to 8, the results are more sensitive to increased vertical resolution than to the addition of stratospheric levels. When the various forecasts are compared with the observations, it is found that for wave numbers greater than or equal to 6, there is no reduction of forecast error from improved vertical resolution. However, for the very long waves in the troposphere there is an error reduction, but of only a few percent after 48 hours.

Simmons (1979a) obtains solutions for forced long waves in a linearized, quasi-geostrophic model which uses sigma-coordinates. The vertical finite difference scheme is the same as the one used in the European Centre for Medium-Range Weather Forecasts (ECMWF) models. These solutions show tropospheric structures are generally less sensitive to stratospheric resolution than indicated in some previous studies. These results were obtained by using stronger wind profiles than those determined by latitudinal averages, and from the inclusion of internal diabatic heating. Steady state solutions are sensitive to stratospheric resolution for cases near resonance; in that case, the slow phase speed of the dominant free mode results in little error over a ten-day period, when accurate initial conditions are specified.

Simmons (1979b) performs a series of forecasts with the standard ECMWF model and another version of this model with poorer resolution in the stratosphere. Both models have 15 levels, but the standard version has the top level at σ=.025 with 4 levels above σ=0.2, while the modified version has the top level at σ=.047 with 3 levels above σ=0.2. A total of 14 cases were run, 8 cases from February 1976, 5 cases from August 1975 and one case from 16 January 1979. The average r.m.s. error of the 500 mb height field is virtually the same for both vertical structures out to 6 days, when the error curves cross the persistence curve. There is a slight improvement for wave numbers 1-3 with the standard vertical structure. Of interest is the forecasts made from 16 January 1979 with both vertical structures, when both models correctly forecast a significant change in the 50 mb height field.

Keeping (1979) compares the long waves in three general circulation models with the following different vertical structures: (1) 5 levels, equally spaced in sigma, (2) 11 levels, extra resolution in boundary layer and in stratosphere, (3) 13 levels, with tropospheric resolution similar to the 5 level model and the highest resolution in the stratosphere. Each model is integrated for at least 50 days and 30-day means are compared with observed means. The long waves generated by the 13-level model were <u>not</u> superior to those produced by the models with poorer stratospheric resolution. There is some indication that the long waves could have been improved by better vertical and horizontal resolution in the troposphere. This improvement could be related to more accurate representations of the forcing effects.

Chao, Schoeberl and Strobel[1] carry out a sensitivity experiment with the UCLA-NEPRF tropospheric forecast model by adding a specified vertical motion at the upper boundary. This vertical motion is present in wavenumbers 1 and 2, and its phase is adjusted to give upward energy flux. Comparisons with a control experiment show a maximum difference of 6 mb at the surface after 3.5 days. However, the authors point out that improved initial conditions would reduce this difference.

Mechoso et al. (1981) compares the following two versions of the latest UCLA general circulation model: (1) 9 levels with the top at 50 mb, (2) 15 levels, first 9 levels the same as model (1), but 6 more levels extending to the top of 1 mb. A sensitivity study is performed by starting both models with the same initial conditions (January) and comparing the solutions. No important differences in the troposphere are observed until after 10 days.

2.4.8 Parameterization of Heating in Stratospheric Models. The diabatic heating of the stratosphere is dominated by the processes of ultraviolet absorption by ozone and infrared emission by carbon dioxide, with lesser contributions from ozone emission and both emission and absorption by water vapor, particularly in the lower stratosphere (see e.g. Houghton, 1978). The parameterization of these radiative processes is complicated by the sensitive dependence of the ozone concentration in the upper stratosphere on temperature, the radiation field, and the presence of trace catalytic chemical species; while in the lower stratosphere the concentration of both ozone and

[1]Chao, W.C., M.R. Schoeberl and D.F. Strobel (1980): Numerical experiments on the stratosphere-tropospheric dynamical interaction. NRL Memorandum Report 4209, Naval Research Laboratory, Washington D.C., 18 pp.

water vapor is largely determined by advection. In stratosphere models the role played by water vapor is generally ignored because it is of secondary importance and is assumed to be handled correctly by parameterization suitable for the troposphere. Thus, the parameterization reduces to describing the radiative heating and cooling due to carbon dioxide and the varying concentrations of ozone.

The infrared cooling rate for a standard atmosphere with an arbitrary ozone concentration profile has been accurately calculated by Dickinson (1973), who gives a corresponding temperature parameterization of the Newtonian cooling rates. This method is now in general use. Schoeberl and Strobel (1978) have improved this parameterization by including photochemical acceleration and adding cooling rates calculated for lower altitudes by Trenberth (1973a). Analytic approximations to the ozone heating rate have been presented by Lacis and Haneen (1974), while Strobel (1976) has developed an alternate parameterization that may be more useful if the incoming solar flux is to be varied.

Using one of the above parameterizations, or detailed radiation calculations if preferred, the diabatic heating of the stratosphere can be closely modeled, if the ozone concentration is known. There are three different approaches to the determination of ozone concentration; it can be arbitrarily specified, separately modeled, or parameterized together with the radiative processes.

Independent specification of the ozone concentration has been used with some success in GCM studies such as those of NCAR (Kasahara _et al._, 1973) and GFDL (Manage and Mahlman, 1976). This approach appears to be particularly applicable when the model only reaches the lower stratosphere, where ozone is a passive tracer (NCAR), or when the specified ozone concentration varies with season and latitude as well as height (GFDL). This method has even been used in models to describe ozone concentration when this feedback is not too important (Harwood and Pyle, 1980). These results suggest a predictive model could successfully use independently specified ozone profiles, if they were regularly updated.

Prediction of ozone concentration in a model requires parameterization of the photochemistry of its creation and destruction. A parameterization of ozone photochemistry has been developed by Cunnold et al. (1975), and this parameterization, together with the separate radiative parameterizations described above, has been incorporated in the UCLA model (Schlesinger and Mintz, 1979) in order to predict ozone concentration in the atmosphere. However, it should be noted that even this parameterization still involves independent specification of the concentrations of the trace catalysts. It is probable that these latter concentrations depend upon solar activity (Heath et al., 1977) as well as terrestrial processes, so there is apparently little hope of being able to predict them. Because of the increased computer resource requirements and the difficulty associated with verification, there does not appear to be much justification for separate detailed calculations of photochemistry outside the context of ozone distribution models.

The possibility of parameterizing the ozone concentration implicitly within the overall radiative parameterization exists because the concentration of ozone in the upper stratosphere is nearly in radiative equilibrim. A technique for accomplishing this parameterization was developed by Leovy (1964), and has found favor in simplified models of the mean stratospheric circulation (Holton and Weinbein, 1980, for example). This approach has also been used successfully in a primitive equation spectral model for the study of sudden stratospheric warming (Lordi, et al., 1980). This method appears to be reliable as long as the excursions of temperature from standard profiles, which are a function of latitude and season, are not too great.

The most satisfactory approach for dynamical modeling would appear to be implicit parameterization of ozone, following the scheme of Schoeberl and Strobel (1978). Of course this scheme can be revised to accommodate newer estimates for the radiative fluxes. A more ambitious program would be to develop a family of parameterizations, e.g. to make use of seasonally monitored ozone profiles, to insure relatively small departures from linearity of the cooling rates. Since stratospheric dynamics do not appear to react back on the troposphere on short time scales, the parameterization of radiative processes should prove adequate. Whether it can be used to predict sudden warmings, however, is not known, but its successful use in models is encouraging.

<u>2.4.9 Stratospheric Data</u>. Routine radiosondes provide observations of the atmosphere up to a height of about 30 km, or roughly the altitude at which radiative effects begin to dominate over the effects of tropospheric motion systems and radiative equilibrium of photochemically active species is established (Holton, 1975). Knowledge of the structure of the stratosphere above this level is based primarily on rocketsondes, but only a few of these stations provide regular coverage. The chief source of data for the stratosphere on a global scale and at regular intervals for initialization and verification must come from remote soundings by satellite-based instruments. For example, infrared radiance data from the Nimbus satellites has been used to monitor temperature fluctuation fields (Stanford, 1979) or available potential energy. One wave mode in the stratosphere has apparently been identified by such techniques (Stanford and Short, 1981). A major difficulty with satellite infrared radiance measurements is the inherent low vertical resolution, but monitoring of long vertical wavelength features can be useful in the stratosphere, since they account for the bulk of the available potential energy (Chen and Stanford, 1980). It is also possible to derive information about small-scale dynamics from high resolution stratospheric ozonometer data (Barat and Aimedien, 1981) which may be useful in developing and improving

parameterization of the turbulent transport. One very promising source of data is the Solar Mesosphere Explorer (SME) mission which will monitor the vertical profiles of temperature, ozone, water vapor and other trace constituents through the stratosphere, from 30 to 80 km (Thomas, et al., 1980). While designed as a one-year experiment, the instrumentation might in principle be used for regular monitoring at some future time. For the present, however, information from the Nimbus series of satellites remains the most complete. Solar occultation techniques may occasionally be used to substantially improve the vertical resolution, but not over large areas on a continuing basis (Russell, 1980). Thus, the vertically integrated infrared radiance profiles remain the principle datum subject to measurement and verification.

SECTION 3 - CONCLUSIONS AND RECOMMENDATIONS

3.1 Inclusion of Stratospheric Modeling. The sudden stratospheric warmings and the other 10-20 day period oscillations in the winter stratosphere include the most important troposphere-stratosphere interactions which could have significant tropospheric effects during medium-range forecasts. Sudden stratospheric warmings are caused by vertically propagating planetary waves which interact with the mean flow. This process can be simulated in a model with forcing at the tropopause and a realistic initial mean wind (subsection 2.4.4). However, full tropospheric-stratospheric models have been mostly unsuccessful in simulating sudden stratospheric warmings (subsection 2.4.5). Linear studies suggest that the planetary waves in the troposphere which are predicted by numerical atmospheric prediction models should be very sensitive to stratospheric resolution (subsection 2.4.6). However, sensitivity studies and forecast intercomparisons have failed to show any significant tropospheric forecast improvement due to added stratospheric resolution (subsection 2.4.6).

The main reason for adding a detailed stratosphere to future FNOC atmospheric forecast models is to improve tropospheric prediction rather than to develop a capability for stratospheric prediction, as indicated by Dr. T. Rosmond. Therefore, it is concluded that the addition of a detailed stratosphere would not be cost effective during the next five years. This conclusion is based on the inability of present models to predict properly significant forcing of the stratosphere, such as sudden stratospheric warmings, and on the lack of tropospheric forecast improvement in various prediction studies.

However, it is recommended that research in this area be carefully monitored, because it is hoped that in the future proper troposphere-stratosphere modeling may allow significant improvements in forecasting changes in blocking patterns and other significant circulation changes in the troposphere. It appears that these large-scale tropospheric features are more dependent on tropospheric forcing. Thus, more research is required to improve the tropospheric diabatic package and the representation of topography effects.

3.2 Coupling of Atmosphere-Ocean Models. Based on the above review and the present state-of-the-art of coupled atmosphere-ocean models, it is concluded that there is inadequate scientific and economic justification for development of a fully synchronous coupled model for operational use during the next five years.

It would be scientifically imprudent to advise operational use of such a coupled model (in the sense of Case C in Exhibit 3) because:

(1) There is a strong likelihood that the north-south sea surface temperature gradient would be distorted by imbalances/incorrect values of surface heat fluxes in equatorial and polar regions. These incorrect values would persist and accumulate in equatorial upwelling regions and polar ice-edge regions, because of a lack of ocean thermal structure observations to constrain these incorrect tendencies. Consequently, a detrimental effect on the "climatology" of the model atmosphere (and ocean) would likely result.

(2) The interaction between cloudiness and insolation in producing east-west SST anomalies in equatorial regions with light surface winds is not sufficiently understood to hazard running a coupled model. Warm and shallow SST anomalies may be produced on time scales of 1-2 days and space scales of about 1000 km. Since even a complete ocean model including currents is unlikely to prevent the accumulation of heat in these shallow layers, there would be an anomalous vertical heat flux to the atmosphere. This would

likely result directly in deep convective heat release over the warm SST anomalies. As has been shown by several studies, equatorial heat fluxes may excite (in less than one week) a response in the middle latitudes as well as in the tropics. The correct prediction of this response is important - but the possibility of specifying incorrectly this response in a coupled model is a great hazard. More basic research and observational studies are needed.

(3) Uncertainties due to errors in the ocean prediction model and its high sensitivity to atmospheric forcing must be carefully studied (and "tuned") over an extended period (perhaps exceeding five years) to develop an operationally useful version. To couple such a model (TOPS in its present state is clearly unsuitable) with an atmospheric model of the complexity of NOGAPS would be ill-advised. One could not easily distinguish the errors introduced in the coupled version, because of the additional degree of freedom. In terms of Exhibit 2, the potential improvements due to the inclusion of time dependent ocean variability on a time scale of 5-10 days are not likely to result consistently in positive impacts greater than the negative impacts.

(4) The potential positive impact on atmospheric prediction to be gained by the inclusion of a coupled ocean model is likely to be smaller than the uncertainties due to improper representation of other physical processes in the model. Chief among these are the vertical fluxes of heat, moisture and momentum due to convective and larger scale atmospheric circulations. One should not assume that our present parameterizations of these physical processes is adequate. Recent studies of maritime

cyclones (see review in Sandgathe, 1981) suggest that these parameterizations must be improved for prediction of explosive cyclogenesis. The development of improved parameterizations will be even more important when the time-dependent sea surface temperatures are available from a coupled atmosphere-ocean model in the future.

(5) Requirements for ocean prediction in support of immediate fleet requirements are likely to be best met by limited domain models. That is, it is likely that high-resolution models will be developed for regions of maximum Naval interest, rather than global models of uniform resolution. Only in these regions with compelling national security implications is it likely that there will be sufficient motivation to acquire the necessary oceanic observations to permit accurate oceanic prediction. It is more likely that the atmospheric forcing for these ocean models will be from regional scale atmospheric models, rather than from NOGAPS. Consequently, a coupling with NOGAPS would not likely be a viable option.

It would not be economically cost effective to develop and run a fully synchronous coupled model because such a model would be:

(1) Extremely costly in terms of computer storage requirements to make a synchronous atmospheric and oceanic model integration. Either model by itself would tax the most advanced computers that will be available during the next five years.

(2) Extremely costly in terms of computer time. The complexity of the required ocean model to produce viable SST forecasts in critical regions, such as

the Kuroshio/Gulf Stream and Equatorial upwelling zones, should not be underestimated. To couple such a model in a synchronous fashion would require reductions in the atmospheric model that would diminish the atmospheric forecast capability more than a coupled model would improve this capability.

(3) Extremely costly in terms of software development and testing. Additional degrees of freedom and coding strategies in the coupled model would have to be developed and tested. Since such research and development is not yet well-advanced, particularly methods of testing such a model in an operational setting, one should proceed systematically at a deliberate pace. This required pace is unlikely to justify a coupled model during the next five years.

It is recommended that the primary emphases during the next five years be two-fold. First, develop an atmospheric model with improved vertical (especially in the lower troposphere) and horizontal resolution to predict properly the response to sea surface temperature gradients in regions of strong air-sea interaction. Further development of parameterizations of the basic atmospheric processes should proceed concurrently. Secondly, develop an oceanic analysis system to produce the best possible representation of sea surface temperature for input to the medium-range atmospheric model. This recommendation is illustrated as Case A of Exhibit 3, in which the sea surface temperature would be held fixed during the atmospheric model integration. The proposed oceanic analysis system does include a short-term oceanic prediction model; however, the purpose of the prediction model is to provide a dynamically consistent evolution of the sea surface temperature that is consistent with the subsurface fields and the recently "observed" surface forcing. That is, the primary objective of this ocean model would be to diagnose the present state of the ocean, and especially establish a representative sea surface temperature. This ocean model should be driven with the best possible representation of the atmospheric forcing. As indicated in Case A, this information would be derived from analyses and short-term predictions as an integral part of the atmospheric four-dimensional data assimilation. The development and operational validation of an improved tropospheric model offers the best potential for improved forecasts of maritime weather systems that impact on Naval operations.

As noted above, the recommended strategy for tropospheric prediction (Case A, Exhibit 3) requires a knowledge of the "observed" SST field at the initial time. If the horizontal and vertical resolution in the tropospheric model is improved as recommended above, the model will become more responsive to the underlying SST. As a result, it will be more essential that the SST analysis be as accurate and representative as possible.

The above review of the state-of-the-art of ocean models (subsection 2.2) suggests that improvements in the SST analysis can be achieved by combining, in an optimal sense, SST fields produced by a number of different sources. These sources should include a <u>satellite-derived SST field</u> (which could be an average over a suitable time period - e.g. 5-10 days), a <u>model (TOPS) predicted SST field</u>, and an analysis based on <u>conventional data</u>. In addition, <u>climatological information</u> (means, variances, space/time correlation scales, etc.) should also be utilized in a modern objective analysis methodology which makes use of the known geographical variations in the SST field and its statistical properties. Since the ocean model predictions of SST are expected to have an important impact on the SST analysis, research and development of ocean prediction models should be encouraged.

It is also recommended that the non-synchronous coupling concept (Case B, Exhibit 3) be tested as the oceanic prediction capability is developed separately by NORDA and other agencies. This type of coupled model can be tested by merely updating the SST field provided the NOGAPS model. With proper attention to the regional and seasonal aspects of the response of the atmospheric to the ocean, the beneficial and harmful aspects of this coupling for atmospheric prediction could be assessed. Such testing would provide a basis for assessing the likely benefits to be gained in the 5-10 year time frame from a coupled atmosphere-ocean model.

3.3 Computer Resources. The current global atmospheric forecast model of FNOC runs on the Cyber 200 model 203 computer and requires approximately 910,000 (decimal) words of central memory, because the whole program must remain in central memory at all times in order to execute within a reasonable amount of time (wall-clock). Though this computer has the capability of virtual memory operation, the I/O data rate is too slow for the volumn of data to be passed to and from the disks in the amount of time available to keep the computations flowing. The present forecast model has 6 levels in the vertical and a horizontal resolution of 2.4°-by-3° (latitude/longitude). The recommended forecast model should have about 12 levels in the vertical and a horizontal resolution of about 2°-by-2° (latitude/longitude). The current FNOC CDC model 203 computer is equipped with 1 million central memory words, with 64 bits of data per word. Obviously, there is no way the present FNOC Cyber model 203 computer can support this recommended new model. The I/O data rate is too slow for this model to execute in a reasonable amount of time in a virtual memory mode and this model will not fit in the available memory. The FNOC Cyber model 203 will accommodate this new forecast model, if the memory is increased to two million words and CDC's (promised) 32-bit Fortran compiler is acquired. If operational requirements dictate that this new model execute in about the same wall-clock time as the present model (a 72 hour forecast in approximately 72 minutes), then the present computer must be upgraded to a model 205 with two million words of memory and the 32-bit Fortran compiler. With this configuration, the new forecast model should execute somewhere between the same wall-clock time as the present model and 1.25 times longer. Thus, a 5 day global atmospheric forecast should take between 2 and 2.5 hours, and a 10 day forecast should consume between 4 and 5 hours of wall-clock time.

Based upon the best information available, the proposed oceanic analysis-forecast system requires a computer with at least two million words and the 32-bit Fortran compiler. Though this system will execute on a model 203 computer (with two million words of memory) a model 205 is required to keep the wall-clock time to a minimum. It is estimated that a global version of TOPS would execute in about 25 to 30 minutes (wall-clock) on the Cyber 205 with two million words of memory and the 32-bit Fortran compiler in a virtual memory mode, which is the same mode TOPS is presently running.

One point which should be clear is that even though the I/O channel speed on the Cyber 205 is 200 M bit/sec (which is four times faster than the Cyber 203), the maximum useful I/O speed to and from the CDC 819 disks is 37.2 M bit/sec. Thus, as long as these disks are utilized by FNOC on their Cyber 200 computer, the virtual memory mode of this computer will be severely limited.

3.4 Manpower Requirements. The manpower requirements presented below are those required to increase the vertical and horizontal resolution of the NOGAPS forecast model to those recommended and to convert the present TOPS model to a global model. Not provided below are the estimates to establish a global Expanded Ocean Thermal-Structure (EOTS) analysis system, which is required to support a global TOPS model. This estimate is not provided because the present and future operational status of EOTS with the Alternating Parallel Analysis (APA) cannot be established and verified with any degree of certainty at the FNOC. However, it has been established that with minimal effort the present Northern Hemisphere EOTS can support the northern portion of the global TOPS model and the southern portion of the global TOPS model may, if required, be supported through Southern Hemisphere climatological "EOTS" fields, which could be established with a little work.

3.4.1 Global Atmospheric Forecast Model. The number of levels and the vertical spatial resolution of the global forecast model in NOGAPS may be increased with a few parameter changes prior to recompilation. Increasing the upper boundary of the model, which is now fixed at 50 mb, so that the upper forecast level will be approximately 50 mb may be handled in a similar manner.

Increasing the horizontal resolution from 2.4°-by-3° to about 2°-by-2° will require from 3 to 6 months effort by a knowledgeable person, such as Dr. T. Rosmond of the Naval Environmental Prediction Research Facility (NEPRF).

In addition to these minimal requirements, the tropospheric research recommended above should involve approximately 3-person-years per year over the next five years.

3.4.2 Global TOPS Model. The Northern Hemisphere TOPS model, which is based upon the 63-by-63 grid for a polar stereographic projection, will require from 6 to 12 months to establish as a global model based upon a latitude/longitude grid. This estimate is based upon a knowledgeable programmer, such as Mr. Mike Clancy of the Naval Ocean Research and Development Activity (NORDA).

APPENDIX A - REFERENCES

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